



RESEARCH LETTER

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Key Points:

- Influence of increase in sea surface temperature resolution in ERA-Interim data set investigated in Gulf Stream region
- Substantial change in atmospheric response consistent with the change in mean sea surface temperature gradient
- Results indicate the significance of resolution in determining the role of oceanic fronts on weather and climate

Supporting Information:

- Supporting Information S1

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The impact of SST resolution change in the ERA-Interim reanalysis on wintertime Gulf Stream frontal air-sea interaction

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Abstract This paper examines the sensitivity to a change in sea surface temperature (SST) resolution of the interaction between atmospheric and oceanic fronts in the Gulf Stream region in the ERA-Interim reanalysis data set. Two periods are considered, January 1979 to December 2001 (SST resolution $1^\circ \times 1^\circ$) and December 2010 to February 2016 (SST resolution $0.05^\circ \times 0.05^\circ$). The winter season from the latter 6 years of high-resolution SST is compared against six random periods of six wintertime seasons from the low-resolution SST period, to assess the robustness of the result against natural climate variability. In all comparisons, a significant change in frontal air-sea sensible heat flux exchange is found that is highly correlated to the change in mean SST gradient. This leads to both increases and decreases in occurrence of atmospheric fronts and mean precipitation of up to 30%. These results reemphasize the importance of high SST resolution in resolving the influence of oceanic fronts on weather and climate.

1. Introduction

The significant role of strong oceanic fronts of sea surface temperature (SST) for variability in the midlatitude climate system is well documented at a variety of different spatial resolutions [e.g., *Chelton et al.*, 2004; *Sampe and Xie*, 2007; *Minobe et al.*, 2008; *Nakamura et al.*, 2008; *Kwon et al.*, 2010; *Booth et al.*, 2012]. Nevertheless, the extent of this role is a question that is still very much under debate. A recent article however has provided evidence to suggest that the key physical mechanism by which these strong oceanic frontal zones exert their influence is through their interaction with sharp fronts of temperature in the atmosphere associated with migrating weather systems [Parfitt et al., 2016]. In that study, a decrease in the magnitude of the Gulf Stream (GS) SST gradient (analogous to a decrease in resolution) resulted in up to 30% less cold fronts being identified across most of the GS front. A mechanism of “thermal damping and strengthening (TDS)” was proposed to explain the observed change (see Figure S1 in the supporting information). These atmospheric fronts are embedded in baroclinic systems that are known to set the time mean atmospheric state in frontal regions such as the GS [O'Neill et al., 2015; Parfitt and Czaja, 2016]. Crucially, not only is TDS highly dependent on the resolution of the SST, but there are also suggestions that there may be a threshold resolution below which this mechanism cannot be resolved [Smirnov et al., 2015; Willison et al., 2013]. Indeed, the underrepresentation of ocean-atmosphere coupling in general has long been linked to underrepresentation of SST gradients [Chelton, 2005]. Given this, it is critical to identify any associated artifacts in reanalysis data sets due to their coarse SST resolution or a change of the SST resolution in time.

Further motivation is provided from recent sensitivity studies to differently resolved Kuroshio Extensions (KE) resulting from the SST changes in the European Centre for Medium-Range Weather Forecasts (ECMWF)-reanalysis (ERA)-Interim data set [Masunaga et al., 2015, 2016]. These series of studies found considerable mesoscale imprints of the KE on the atmospheric boundary layer that were present for the reanalysis only where the prescribed SST resolution was high enough. The aim of this article is to investigate the impact of this changing SST resolution in the ERA-Interim data set on the interaction between oceanic and atmospheric fronts. The focus will be on the wintertime GS sector, an area of substantial SST gradient, and the region where the frequency of midlatitude atmospheric fronts is the largest [Berry et al., 2011a]. In section 2, the data and methods used are presented. The results are presented in section 3, and a summary and discussion are provided in section 4.

2. Data and Method

ERA-Interim is a global atmospheric reanalysis data set [Simmons *et al.*, 2007; Berrisford *et al.*, 2009; Dee *et al.*, 2011] available from January 1979 to present, with spectral resolution T255 ($\sim 0.7^\circ$) available on a $0.75^\circ \times 0.75^\circ$ longitude-latitude grid. In this study, 12-hourly fields (at 0000 UTC and 1200 UTC) of the horizontal winds and air temperature are used at a pressure level of 925 hPa. Also used in this study are the sensible heat flux, convective precipitation, and large-scale precipitation that are taken from short-range forecast accumulations at both 0000 UTC and 1200 UTC.

However, despite the atmospheric model resolution remaining constant throughout the whole data set, the prescribed SST data have been improved twice. Between January 1979 and December 2001 the SST resolution is $1.0^\circ \times 1.0^\circ$, whereas between January 2002 and January 2009 it is $0.5^\circ \times 0.5^\circ$. From February 2009 onward, the SST resolution has been $0.05^\circ \times 0.05^\circ$. This study seeks to assess sensitivity to SST resolution by comparing the 6 year wintertime period DJF (December–February, where the year denotes the December month—i.e., DJF2010 is December 2010, January 2011, and February 2011) 2010–2015 (“High-res” SST resolution $0.05^\circ \times 0.05^\circ$) against six randomly selected 6 year DJF wintertime seasons within the period January 1979 to December 2001 (“Low-res” SST resolution $1.0^\circ \times 1.0^\circ$). The comparison is repeated using the six different low SST resolution periods to assess the robustness of results independent of the phase of the natural climate variability, e.g., North Atlantic Oscillation. A summary is provided in Table 1. Figure 1a illustrates the difference in mean SST for Low-res sample i , $\overline{\text{SST}}_{(i)} = \overline{\text{SST}}_{\text{low-res}(i)} - \overline{\text{SST}}_{\text{high-res}}$. The plots for the other Low-res samples are shown in the supporting information Figure S1. Although subtle differences exist between each of these plots, the dominant signal ($\Delta \text{SST} \sim \pm 2\text{K}$) following the GS front is highly consistent among all six of them. Figure 1b illustrates the corresponding difference in the magnitude of the mean SST gradient, $|\overline{\nabla \text{SST}}_{\text{low-res}(i)}| - |\overline{\nabla \text{SST}}_{\text{high-res}}|$ for Low-res sample i . Again, the plots for the other Low-res samples are shown in supporting information Figure S1. As before, the magnitude and pattern of the differences are extremely similar for the six different comparisons. This illustrates a sizable permanent artifact in both the SST and the SST gradient that results from the change in the SST resolution regardless of the phase of the internal climate variability.

In order to assess the impact of this artifact on the regional frontal air-sea interaction, atmospheric frontal grid points are identified with the “ F diagnostic” [Sheldon, 2015; Parfitt *et al.*, 2016]. Regions where $F = \frac{\zeta_{925} |\nabla T_{925}|}{\zeta_0 |\nabla T_0|} > 1$ are those considered to be frontal grid points, where $|\nabla T_{925}|$ is the magnitude of the temperature gradient on the 925 hPa pressure level, ζ_{925} is the isobaric relative vorticity on that same pressure surface, ζ_0 is the Coriolis parameter at 40°N , and $|\nabla T_0|$ is a typical scale for temperature gradient, 1 K/100 km. Cold fronts are identified via $\mathbf{u}_{925} \cdot \nabla T_{925} > 0$, where \mathbf{u}_{925} is the horizontal velocity at 925 hPa, as is commonly suggested [e.g., Hewson, 1998]. It is noted that a change in the F threshold preferentially selects stronger or weaker frontal regions but does not change the conclusions reached in this study.

3. Results

3.1. Change in Frequency of Cold Fronts

Figures 2a and 2b illustrate the frequency of all atmospheric fronts and cold fronts for the High-res period DJF 2010–2015, respectively, as a fraction of the total 6 year wintertime period. As can be seen from these two figures, the cold frontal frequency is the main contributor to the total frontal frequency. Indeed, this is as expected since other studies have shown the cold frontal frequency in the wintertime Gulf Stream region to be roughly twice as frequent as warm fronts [e.g., Berry *et al.*, 2011a]. For this reason, the remainder of the study focuses on cold fronts exclusively. Figure 2c depicts the cold frontal frequency for Low-res sample i , while the same plots for the other Low-res samples are shown in the supporting information Figure S3. For each sample, there is a broad frequency signal across the GS front of $\sim 10\%$ of the respective total 6 year wintertime period. However, there are subtle differences in both the magnitude and spatial structure, as one might expect from large-scale variability between each sample.

Figure 2d illustrates the percentage change in cold frontal frequency between the High-res period and Low-res sample i (i.e., Low-res sample–High res period, such that blue (red) indicates a higher (lower) frontal frequency in the High-res period). Corresponding plots for the other Low-res samples are shown in the

Table 1. A Summary of the Six Randomly Selected Wintertime Seasons From the Low SST Resolution Period and the 6 Year Wintertime High SST Resolution Period^a

Samples	Years (for DJF Data)	(I) Correlation Coefficient	(II) Correlation Coefficient	(III) Correlation Coefficient
Low-res <i>i</i>	1983, 1985, 1986, 1987, 1992, and 1998		0.784	
Low-res <i>ii</i>	1979, 1980, 1981, 1983, 1990, and 1999	0.605	0.795	0.723
Low-res <i>iii</i>	1987, 1988, 1990, 1994, 1995, and 1996	0.690	0.791	0.655
Low-res <i>iv</i>	1981, 1985, 1987, 1988, 1997, and 1998	0.829	0.781	0.857
Low-res <i>v</i>	1980, 1989, 1990, 1992, 1993, and 1999	0.584	0.780	0.759
Low-res <i>vi</i>	1979, 1983, 1984, 1992, 1994, and 1997	0.703	0.757	0.784
High-res	2010, 2011, 2012, 2013, 2014, and 2015			

^a(I, II, and III) Spatial pattern correlations for the three different pairs of variables in the domain (28.5°–78°W, 31.5°–52.5°N). (I) The correlation coefficients between the frequency change for the Low-res sample *i* (Figure 2d) and the same quantity for the other Low-res samples (Figures S3f–S3j). (II) The correlation coefficients between the change in dQ/dy between each Low-res sample with the High-res period (Figures 3c and S4f–S4j) and the corresponding change in $|\nabla \text{SST}_{\text{low-res}(i)}|$ (Figures 1b and S2f–S2j). (III) The correlation coefficients between the change in total precipitation for the Low-res sample *i* (Figure 4c) and the same quantity for the other Low-res samples (Figures S7f–S7j).

supporting information Figure S3. For all Low-res samples there is a strong decrease across the majority of the GS front, but with increases to the north and south, revealing the presence of a tripole pattern of change in the wider GS region. In each case, the magnitude of change present in each tripole branch is roughly the same, with absolute values up to 30%. This consistency between the six different comparisons demonstrates the robustness of the signal to natural large-scale variability. In fact, the spatial correlations

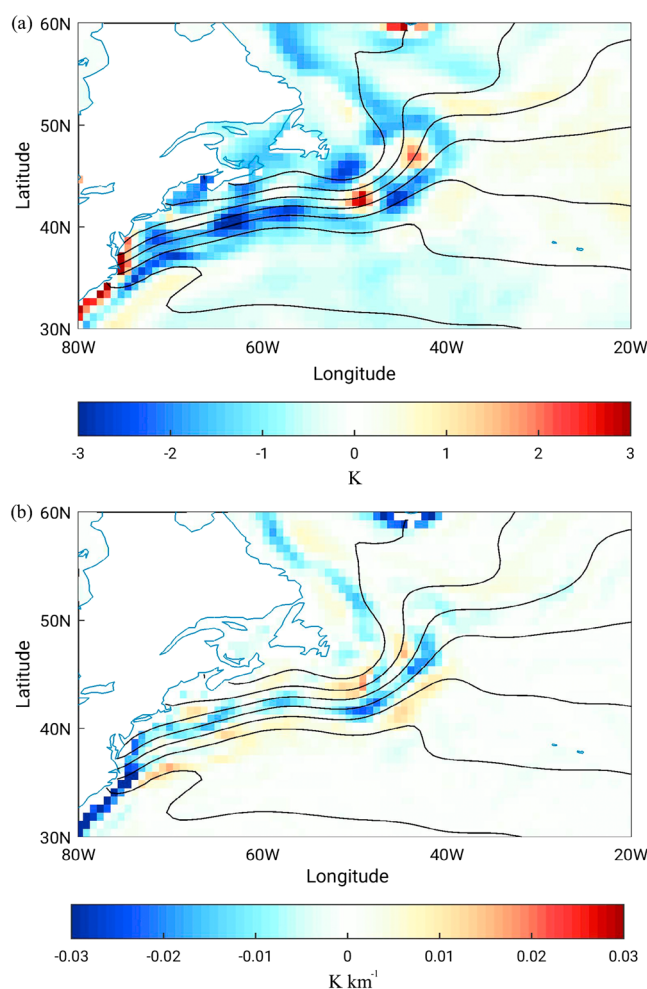


Figure 1. (a) The difference in mean SST between the High-res period and Low-res sample *i*, $\overline{\text{SST}_{\text{low-res}(i)}} - \overline{\text{SST}_{\text{high-res}}}$. (b) The corresponding difference in the magnitude of the mean SST gradient, $|\overline{\text{SST}_{\text{low-res}(i)}}| - |\overline{\text{SST}_{\text{high-res}}}|$. Mean SST contours for Low-res sample *i* are plotted from 3°C to 24°C, at 3°C intervals.

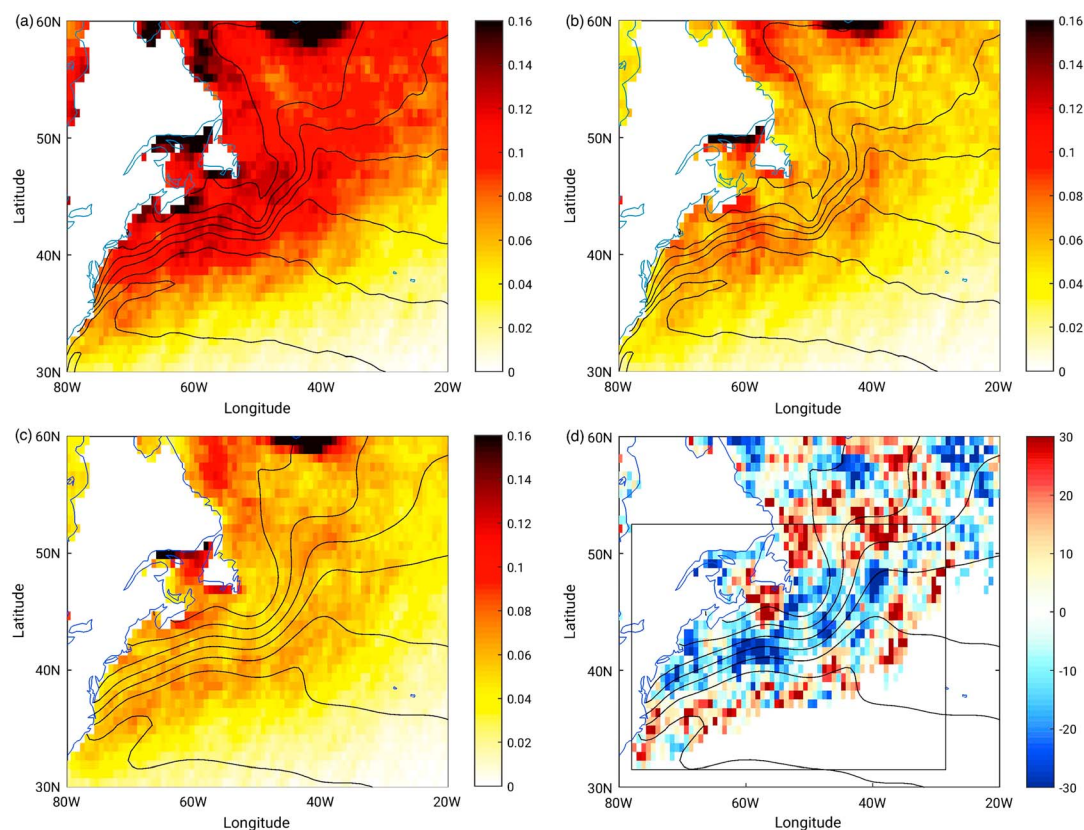


Figure 2. The frequency of (a) all atmospheric fronts and (b) cold fronts only, for the High-res period, as a fraction of the total 6 year High-res period. Mean SST contours for the High-res period are plotted from 3°C to 24°C, at 3°C intervals. (c) The cold frontal frequency for Low-res sample *i*. (d) The percentage change in cold frontal frequency between the High-res period and Low-res sample *i*. The difference is calculated as the Low-res sample minus the High-res period, such that blue (red) indicates a higher (lower) frontal frequency in the High-res period. For Figures 2c and 2d, mean SST contours for Low-res sample *i* are plotted from 3°C to 24°C, at 3°C intervals. The rectangular black box in Figure 2d indicates the domain (28.5°–78°W, 31.5°–52.5°N).

between the percentage change in a GS rectangular domain (28.5°–78°W, 31.5°–52.5°N), for Low-res sample *i*, plotted in Figure 2d, and the percentage change in the same domain in each of the other Low-res samples show strikingly high values (Table 1(l)).

The broad decrease in cold frontal frequency of up to ~30% across the majority of the GS front is of a similar magnitude to that found in the previous study discussed in section 1 that investigated the sensitivity of cold frontal frequency to changing SST resolution in an atmospheric general circulation model (AGCM) across the GS region [Parfitt *et al.*, 2016]. The TDS mechanism, to which the changes in the AGCM experiments was attributed, is heavily tied to the how well the SST *gradient* is resolved, which primarily affects the strength of the air-sea sensible heat flux gradient across atmospheric fronts passing through the region. The next section addresses this issue further.

3.2. Change in Cross Cold-Frontal Sensible Heat Flux Gradient

Figure 3b illustrates the mean cross-frontal air-sea sensible heat flux gradient, dQ/dy (in $\text{Wm}^{-2}/100\text{ km}$, where Q is defined as positive when heat is released from the ocean into the atmosphere), experienced by cold atmospheric fronts at each location for the High-Res period DJF 2010–2015. More specifically, whenever a cold atmospheric front is identified at a location, a vector (y) that points perpendicularly toward the cold sector is established. The sensible heat flux gradient along that vector (across the front) is then calculated, and a composite mean at each location across the period is calculated. Because the gradient is defined positively toward the cold sector, wherever dQ/dy is positive (negative), the sensible heat flux gradient on average acts to thermally dampen (strengthen) the cold front, via TDS (supporting information Figure S1).

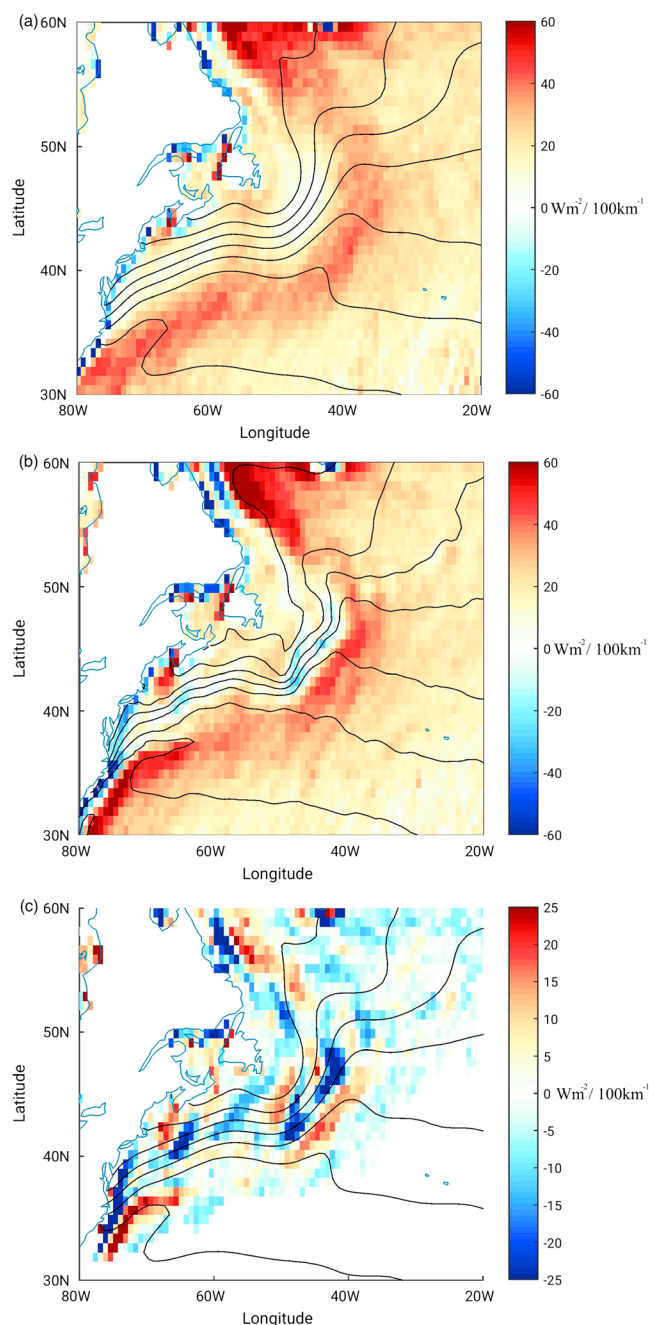


Figure 3. The average cross-frontal air-sea sensible heat flux gradient, dQ/dy (defined in the text), experienced by cold atmospheric fronts at each location for (a) Low-Res sample i and (b) the High-Res period. (c) The difference in magnitude between the High-res period and the 6 year Low-res sample i , calculated as the High-res period minus the Low-res samples (note that the sign convention is opposite to Figures 1 and 2), such that a negative (positive) difference implies that atmospheric cold fronts are on average dampened more (less) in the Low-res sample than in the High-res period. The mean SST contours from Low-res sample i are plotted from 3°C to 24°C , at 3°C intervals in Figures 3a and 3c, and those from the High-res period are in Figure 3b.

with the plots for the other Low-res samples shown in the supporting information Figure S4. Where this difference is negative (positive), atmospheric cold fronts are on average dampened more (less) in the Low-res sample than in the High-res period. Just off the continent exists a region of negative difference

While in the region within the GS box but away from the GS front, the air-sea heat exchange acts to slightly dampen the cold fronts (~ 20 to $30 \text{ Wm}^{-2}/100 \text{ km}$), a dipole exists across the GS front. To the north of the front there is a region where the air-sea heat exchange acts neutrally ($\sim 0 \text{ Wm}^{-2}/100 \text{ km}$) or to strengthen the cold fronts (~ -10 to $-15 \text{ Wm}^{-2}/100 \text{ km}$). To the south of the front however, there is a region of enhanced dampening ($>40 \text{ Wm}^{-2}/100 \text{ km}$). Such a distribution is strongly preferential to a pattern of intensification of passing atmospheric cold fronts that is aligned with the strong GS front. Figure 3a illustrates the analogous composite to Figure 3b but for Low-res sample i , with those for the remaining Low-res samples shown in the supporting information Figure S4. As in Figure 3a, in each Low-res sample for the region within the GS box but away from the GS front the tendency is for the air-sea heat exchange to moderately dampen the cold fronts (~ 20 – $30 \text{ Wm}^{-2}/100 \text{ km}$), while a dipolar pattern exists across the GS front. However, there no longer exists a region along the GS front where the air-sea heat exchange acts neutrally or to strengthen the cold fronts. While to the south of the front there still exists a region of enhanced dampening ($>40 \text{ Wm}^{-2}/100 \text{ km}$), to the north there is now simply a region of weaker dampening ($\sim 10 \text{ Wm}^{-2}/100 \text{ km}$). In other words, a less resolved GS front on average acts everywhere to dampen passing atmospheric cold fronts through air-sea sensible heat flux exchange. This phenomenon is consistent with that found in Parfitt *et al.* [2016].

Figure 3c shows the difference in magnitude between the High-res period and Low-res sample i (i.e., High-res period–Low-res sample),

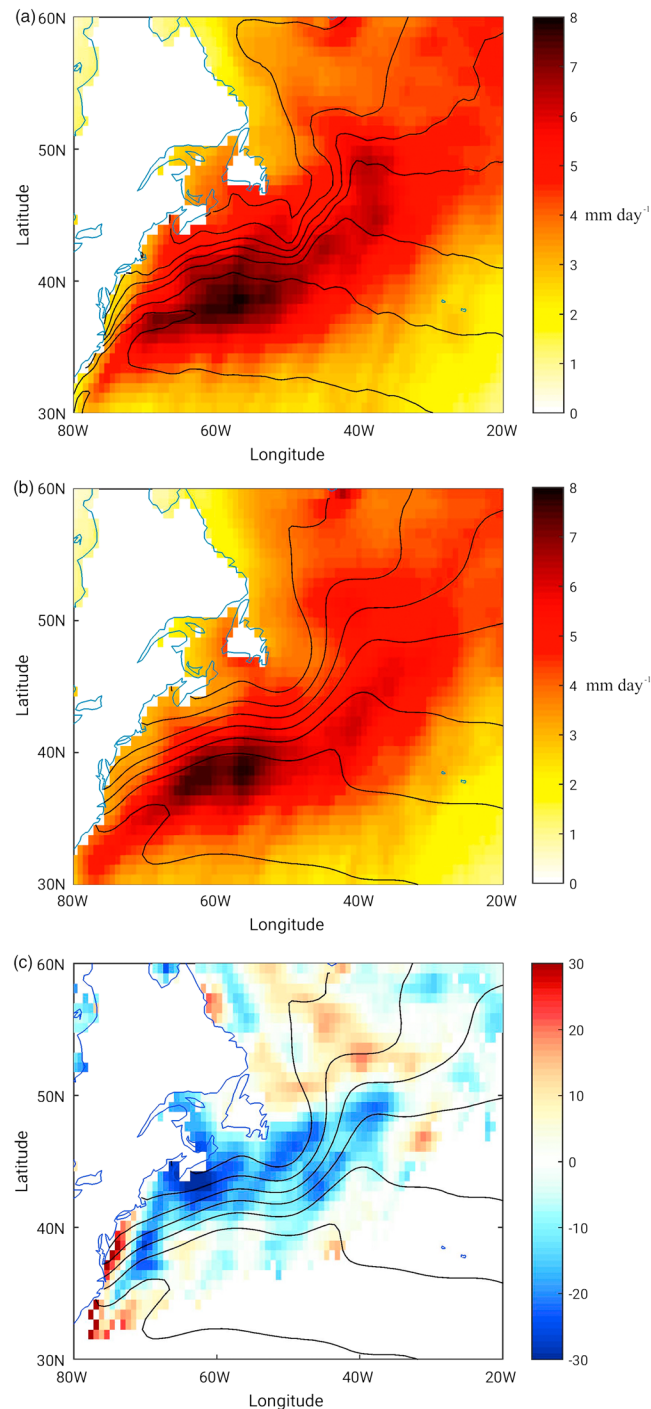


Figure 4. The mean total precipitation for (a) the High-res period and (b) Low-res sample *i*. (c) The percentage difference between the High-res period and Low-res sample *i*. This difference is calculated as $100 \times (\text{Low res} - \text{High res}) / (\text{High res})$ such that negative (positive) values imply more (less) precipitation in the High-res period. The mean SST contours from Low-res sample *i* are plotted from 3°C to 24°C, at 3°C intervals, in Figures 4b and 4c. The mean SST contours for the High-res period are plotted in Figure 4a.

that meanders with the GS front along the extent of the coast. All atmospheric cold fronts propagating off the continent across this region will therefore experience a stronger dampening in the Low-res samples than in the High-res period. As these storms travel on the order of 10 m s^{-1} [Neu et al., 2013], any thermal frontal air-sea interaction on a time scale of up to 1 day can potentially translate into a difference of up to roughly 10° . The slightly shifted spatial location and slightly broader region of cold frontal frequency decrease between 38° and 50°N shown in Figure 2d is thus not unexpected. This shifted spatial relationship is shown explicitly in the supporting information Figure S5. A corresponding shifted spatial relationship between weaker dampening and increased frontal frequency in the Low-res samples is also shown in the supporting information Figure S6. Outside of these regions this consistency is harder to highlight visually due to the complex structure of the change in dQ/dy . However, a spatial correlation in the GS rectangular domain (28.5° – 78°W , 31.5° – 52.5°N) of the difference in dQ/dy between each Low-res sample with the High-res period and the associated difference in $|\overline{\nabla \text{SST}}_{\text{low-res}(n)}|$ reveals a very high correlation (Table 1(II)). In other words, the changes in the gradient of air-sea sensible heat flux across atmospheric cold fronts that appear to be modulating the changes in the atmospheric cold frontal frequency in the region can be mostly explained by the change in the mean SST gradient between the Low-res samples and the High-res period.

3.3. Change in Total Precipitation

The observed impact of the change in SST gradient on atmospheric cold frontal frequency will naturally affect associated variables. An obvious

example is the regional precipitation; in the Gulf Stream region extreme precipitation events are heavily biased toward frontal systems [Parfitt and Czaja, 2016]. Figure 4a illustrates the mean total precipitation

(convective and large scale) for the High-res period in mm d^{-1} , whereas Figure 4b shows the mean total precipitation for Low-res sample i . The mean total precipitation for each other Low-res sample is shown in supporting information Figure S7. While there is a general spatial structure in each Low-res sample that meanders with the GS front, there are subtle differences in both distribution and magnitude between them as one would expect from natural variability. However, Figure 4c, which illustrates the percentage difference between the High-res period and Low-res sample i shows that there is a broad decrease of up to 30% in total precipitation across the entire GS front, a signal which is echoed in each Low-res sample (supporting information Figure S7). It is noted that this percentage decrease, in terms of magnitude and general spatial structure, is seen almost equally in both the convective and large-scale precipitation separately (not shown). Indeed, the magnitude of this decrease in total precipitation is nearly the same as the broad decrease in atmospheric cold fronts observed in Figure 2 across the GS front. This relationship is perhaps not entirely unexpected, given that up to 90% of the total precipitation is associated with extratropical cyclones [Hawcroft *et al.*, 2012]. Once again, the consistency between the six different comparisons illustrates the robustness of this signal to large-scale variability (Table 1(III)). It is emphasized that to the north and south of this broad region of decrease are regions where the precipitation actually increases, consistent with the tripole pattern of cold frontal frequency change observed in Figure 2.

4. Summary and Discussion

This article has examined how sensitive the interaction between atmospheric and oceanic fronts in the wintertime Gulf Stream region is to SST resolution in the ERA-Interim reanalysis data set. Six random selections of six winter seasons (DJF) were selected from the Low-res period January 1979 to December 2001 and compared with 6 years from a High-res period DJF 2010–2015. In all six comparisons, the frequency of atmospheric cold fronts was reduced by up to 30% in the Low-res period across the GS front. To the north and south of the strongest SST gradient, an increase of up to 30% was present. The high correlation between the changes in each six comparisons indicates the robustness of the signal to natural variability. These magnitudes of change are also noticeably larger than those expected from any long-term trend [e.g., Berry *et al.*, 2011b]. Subsequent analysis of the changes in air-sea sensible heat flux gradient across those cold fronts revealed the frequency change to be consistent with the mechanism thermal damping and strengthening [Parfitt *et al.*, 2016]. This mechanism is highly dependent on how well both the atmospheric and oceanic fronts are resolved, and a spatial analysis between the changes in air-sea sensible heat flux gradient and the associated changes in SST gradient indeed revealed that both are highly correlated. It is duly noted however that other factors, such as the impact of the resolution change on vertical winds within the boundary layer, may also contribute to the magnitude and pattern of the frontal frequency change. As one might expect, the changes in the frequency of cold fronts are accompanied by corresponding changes in the surface storm-track activity (supporting information Figure S8, based on the definition of Wallace *et al.* [1988]). These results, along with other recent studies [e.g., Parfitt *et al.*, 2016], suggest that the impact of the SST gradient could potentially be stronger than that of the absolute SST in modulating passing storms, although the role of SST is still highly important [Booth *et al.*, 2012]. Furthermore, robust changes of up to 30% are also found in each of the six comparisons for the total precipitation, which are consistent with the structure and magnitude of change in the cold frontal frequency. This is to be expected given the high correlation between atmospheric cold fronts and precipitation along the midlatitude storm tracks, especially in the wintertime Gulf Stream region [Catto *et al.*, 2012].

All of the results in this study were reproduced for six additional low-resolution periods in which each period contained consecutive (rather than random) years, as well as 20 more distinct periods of six random years (not shown); in each case, the conclusions stay robust. It is noted that the 6 year High-res period has an average December–March North Atlantic Oscillation index [Hurrell, 1995] that is roughly the same as the median for the distribution of every possible 6 year Low-res sample, which minimizes any influence of large-scale interannual variability on these results (supporting information Figure S9).

Several recent studies have further hinted at the importance of SST gradient in the role that oceanic fronts play in affecting midlatitude weather and climate. For example, Kwon and Joyce [2013] showed that decadal variability in the Gulf Stream path has a significant effect on large-scale atmosphere variability, with associated anomalous patterns in SST gradient. Additionally, Parfitt [2014] illustrated that a change in the Gulf Stream SST gradient resulted in a “steering” of passing atmospheric fronts that noticeably changed

their direction of propagation by up to 10°. The results shown in this study further suggest an important role for the SST gradient in how oceanic fronts affect midlatitude weather and climate. One avenue currently being explored is the possibility of developing an oceanic index for western boundary currents specifically based on the SST gradient (e.g., the Oyashio Extension index used by *Frankignoul et al.* [2011] and *Kwon and Joyce* [2013]) or the surface heat flux exchange gradient, as opposed to other more traditionally used variables such as the sea surface height [e.g., *Qiu et al.*, 2014].

It is noted again that just as crucial as the SST gradient is the resolution of the SST data itself. Both are intimately connected as any change in the resolution will automatically affect the gradient. Given the magnitude of the artifacts in frontal frequency and precipitation observed in this study as a result of the change in SST resolution (and subsequently changes in SST gradient), caution must be taken when analyzing long-term trends in either reanalysis or model data sets within which the SST resolution changes or in making comparisons between two data sets in which the SST resolution is different. Indeed, it is emphasized that these artifacts exist in the ERA-Interim reanalysis despite the presence of data assimilation, meaning that the impact of the increase in SST resolution is rising above any assimilation constraints. Indeed, if the constraints were completely sufficient, one would expect to solely observe interannual variability. Furthermore, given recent studies that have suggested the interaction between atmospheric and oceanic fronts can only be properly considered at sufficiently high resolution [*Smirnov et al.*, 2015; *Parfitt et al.*, 2016], it is likely that an increasing impact of oceanic frontal zones on large-scale variability will be observed as the resolution of climate models increases. The exact magnitude, extent, and mechanism of this increase however is still an open question, and further research is presently being conducted by the authors using both high- and low-resolution coupled climate model simulations.

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